
Interpreting the Dimensions of Ancient Fluvial Channel Bars, Channels, and Channel Belts from Wireline-Logs and Cores¹

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ABSTRACT

A primary objective in exploration for and development of fluvial reservoirs is determining the thickness and width of sandstone-conglomerate bodies (mainly channel-belt deposits). Most of the existing techniques for estimating the dimensions of fluvial reservoirs have major drawbacks. A fresh approach to the problem is made using recent theoretical, experimental, and field studies. This new approach involves (1) new models for the lateral and vertical variation of lithofacies and petrophysical-log response of river-channel deposits with explicit recognition of the different superimposed scales of strata, (2) distinction among single and superimposed channel bars, channels, and channel belts, (3) interpretation of maximum paleochannel depth from the thickness of channel bars and the thickness of sets of cross-strata formed by dunes, and (4) evaluation of various methods for estimation of widths of sandstone-conglomerate bodies that represent either single or connected channel belts (outcrop analogs; correlation of sandstone-conglomerate bodies between wells; use of empirical equations relating channel depth, channel width, and channel-belt width; theoretical models; and three-dimensional seismic data).

Two fluvial reservoirs were reinterpreted using this new approach. In the first example from the Mesaverde Group, Colorado, maximum paleochannel depth had been underestimated because the degree of superposition of channel bars had been overestimated. As a result, channel-belt widths determined from empirical equations were underestimated. In the second example from the Travis

Peak Formation, Texas, channel-belt width and connectivity of channel-belt sandstone bodies had been overestimated because of overzealous well-to-well correlation and inappropriate use of width and thickness data from supposed analogs. These examples demonstrate the potential value of this new approach in reservoir characterization and management.

INTRODUCTION

Fluvial sandstone-conglomerate bodies are reservoirs for oil and gas in many of the important hydrocarbon provinces of the world (Atkinson et al., 1990; Halbouty et al., 1970; Kerr et al., 1999; Lawton et al., 1987; Stäuble and Milius, 1970; Swanson, 1993); moreover, much of the world's potable water is stored in aquifers of fluvial origin. To determine reservoir volume and producibility, quantitative estimates are required of the lithofacies (calibrated to porosity and permeability), geometry, orientation, spatial distribution, proportion, and connectedness of permeable and impermeable rock bodies (Bryant and Flint, 1993). Such geologic data are essential to the building and conditioning of three-dimensional (3-D) reservoir models upon which field-development and well-completion strategies are based. Data available to address subsurface modeling needs typically constitute wireline-logs, cores, cuttings, and seismic. Unfortunately, the most desirable data, cores and high-quality 3-D seismic, are commonly sparse or unavailable; therefore, geologic reservoir descriptions are commonly biased or inaccurate, owing to reliance upon wireline-log data.

A primary exploration and development objective is the determination of the width of sandstone-conglomerate bodies in various directions. The sandstone-conglomerate bodies of interest are normally channel belts or connected channel belts (Figure 1). Several techniques are employed to determine their widths: (1) correlation of log signatures between wells, (2) use of outcrop analogs, (3) empirical equations derived from studies of modern rivers

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relating channel-bar thickness (determined from wireline-logs and cores) to maximum paleochannel depth, channel width, and channel-belt width (hence width of sandstone-conglomerate bodies normal to paleoflow direction), and (4) direct estimation from amplitude analysis of 3-D seismic time slices. Each technique suffers from major drawbacks, as discussed in following sections of this paper; nevertheless, the reliability of any of these techniques depends upon correctly interpreting depositional settings. Of special importance is knowledge of paleochannel geometry and mode of paleochannel movement. Unfortunately, our understanding of fluvial depositional processes is incomplete, and most published models of fluvial deposits (Galloway and Hobday, 1996; Miall, 1992, 1996; Selley, 1996) are of limited use in interpreting depositional setting because they are qualitative, lack detail, are not fully 3-D, and are commonly misleading (Bridge, 1985, 1993; Brierley and Hickin, 1991).

Recent studies of modern fluvial processes of deposition, in combination with theoretical modeling, make it possible to approach this subsurface-interpretation problem with a clearer vision. The purpose of this paper is to (1) present models for the vertical and lateral variation of lithofacies and wireline-log response of river-channel deposits, (2) provide ways of distinguishing single and superimposed channels and channel belts, (3) describe a new method for interpreting the spatial variation of maximum bank-full-channel depth from core and wireline-log data, and (4) discuss methods for estimating width of sandstone-conglomerate bodies representing either single or connected channel belts.

MODELS FOR VERTICAL AND LATERAL VARIATION OF LITHOFACIES AND WIRELINE-LOG RESPONSE IN RIVER-CHANNEL DEPOSITS

Single Channel Belts

Different Scales of Deposition in Channel Belts

Simplified plan forms, cross profiles in various positions and orientations, and large-scale stratal geometry of channel belts for braided and unbraided, sinuous rivers are shown in Figure 2. The modes of channel movement implied in Figure 2 partly determine the preservation of channel bars and channel fills, and they are therefore an important component of fluvial depositional models. Idealized vertical sequences of lithofacies and wireline-log response at various locations within channel belts are shown in Figure 3. These diagrams are based on a large amount of research on modern river deposition (see review in Bridge, 1993; also Bernard et al., 1970; Bridge et al., 1995, 1998; Jordan and Pryor, 1992). Four scales of deposition are clearly

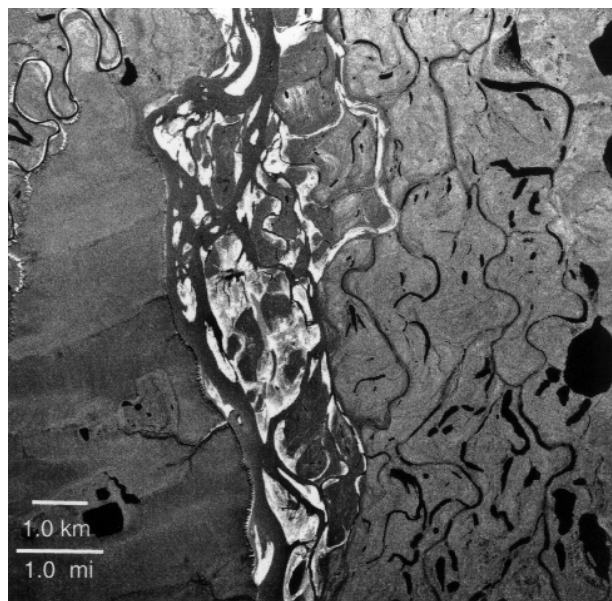


Figure 1—Three channel belts present in part of the Colville River flood plain, Alaska. Note the juxtaposition of the narrow channel belt of the Kogosukruk River (left) formed by a single, sinuous channel and the multiple, sinuous-to-braided channels in the much wider Colville River channel belt (center). Active and abandoned channels and bars are easily discernible. To the right is another abandoned channel belt. Each of these channel belts could be identified in the subsurface using core and wireline-logs, and accurate predictions of their dimensions could be made. Photograph from July 1979 in the National Petroleum Reserve Alaska, approximately 40 km (25 mi) northeast of Umiat.

discernible from Figures 2 and 3: (1) a complete channel belt (the entire sandstone-conglomerate body), (2) deposits of individual channel bars and channel fills (sets of large-scale inclined strata, also known as storeys), (3) sedimentation units (large-scale inclined strata) on channel bars and in channel fills formed by discrete episodes of flooding, and (4) sedimentation units (sets of small- and medium-scale cross-strata and planar strata) created by the passage of bed waves, such as ripples, dunes, and bed-load sheets. These four scales of deposit should be discernible in cores and in image logs. The three largest scales should be discernible from high-quality wireline-logs (Figure 3); furthermore, thickness ratios calculated for the different scales of strata set should be understandable and predictable.

Channel-Bar Deposits (Sets of Large-Scale Inclined Strata)

The geometry and mode of migration of river channels and associated bars have an important

bearing on the geometry and orientation of large-scale inclined strata and on the preservability of strata from different locations in channels and bars. Large-scale strata set (story) thickness in a single channel belt can vary laterally by a factor of two or more (Figure 2). In places, large-scale strata sets thicken laterally as the large-scale strata increase in inclination. Some sets have large-scale strata inclined predominantly in one direction, whereas other sets show convex-upward or concave-upward stratal inclinations. By recognizing these patterns of large-scale strata in cross section, deposits of braided and unbraided rivers and of channels of differing sinuosity can be distinguished (Bridge, 1985, 1993). It follows that channel patterns (e.g., braided, meandering) cannot be interpreted from vertical lithofacies profiles, contrary to published opinions (Collinson, 1996; Galloway and Hobday, 1996; Miall, 1992, 1996; Selley, 1996). It should also be appreciated, when attempting to reconstruct paleochannel patterns from ancient deposits, that channel patterns in a particular reach of a channel belt can vary markedly in space and time. This may be due, for example, to local variations in bank materials, localized tectonism, the effects of particularly severe floods, or bend cutoffs.

The number of large-scale inclined strata constituting most of the thickness of a set (story) is commonly between 1 and 10 (Willis, 1993b; Bridge et al., 2000) (four or five large-scale inclined strata constitute most of the thickness of the stories in Figure 3). This number depends on the rate of lateral migration relative to channel-bar width as measured in a section of a given orientation. For example, if the channel migrates a distance equivalent to the apparent bar width during a single depositional event, the bar sequence will constitute a single large-scale stratum. If 10 depositional episodes are required to migrate one bar width, 10 large-scale strata will be formed. The amount of channel migration during a flood is commonly on the order of $10^{-1} \times$ channel width. In cores or wireline-logs, strata interpreted as large-scale inclined strata could be confused with individual channel-bar sequences (Figure 3), leading to gross underestimates of paleochannel depth; therefore, this approximate relationship between the thickness of large-scale inclined strata and large-scale strata sets should be useful, as illustrated in the following paragraphs in our reevaluation of previous interpretations of certain fluvial sandstone bodies.

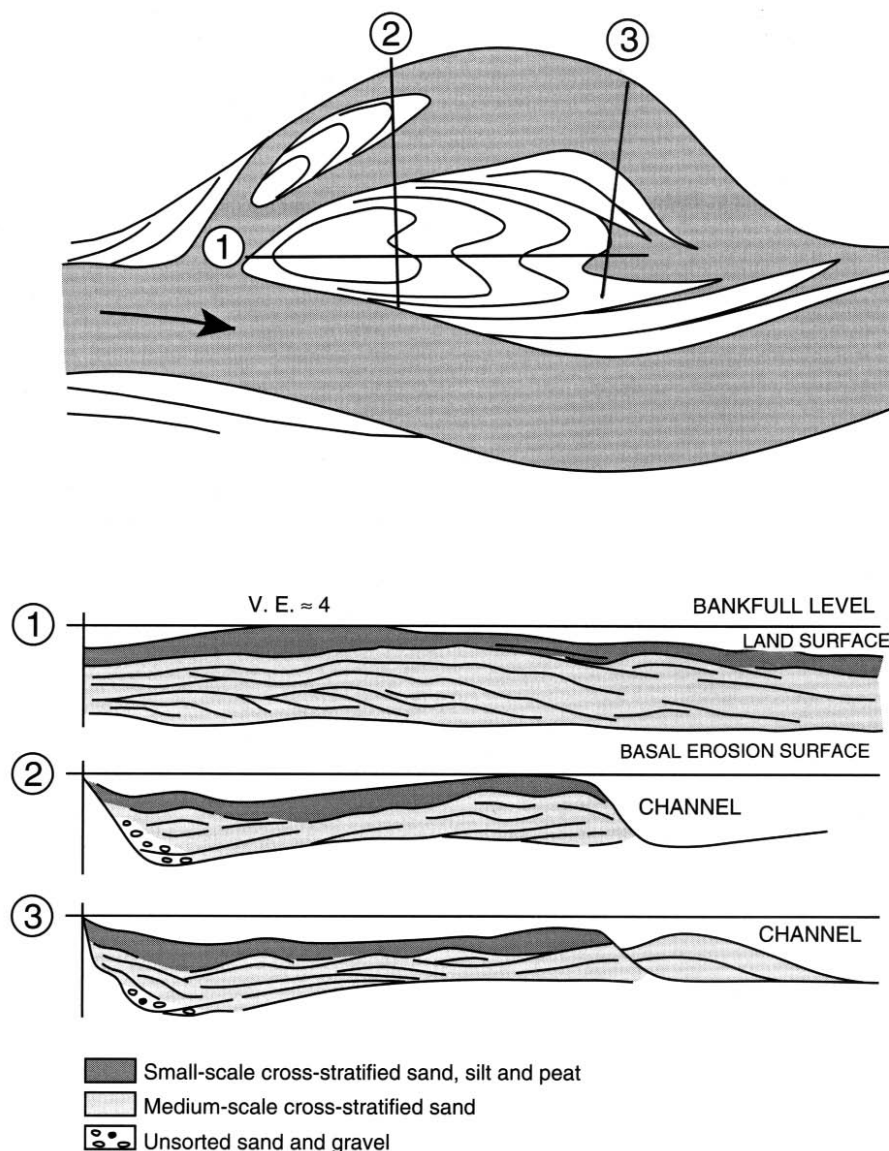
Many river channels and associated bars migrate by downstream translation and lateral expansion (increasing sinuosity and amplitude) of curved channel segments (Figure 2) (Bridge, 1993; Jackson, 1976a; Willis, 1989, 1993a, b). If downstream translation is dominant, only the downstream parts of channel bars can be preserved. Downstream parts

of channel bars tend to exhibit fining-upward profiles. As lateral expansion of channels becomes important, it is possible to preserve more of the upstream parts of bars. These tend to have little vertical variation in mean grain size and, in places, upstream parts of bars may coarsen upward (e.g., Bluck, 1971; Bridge and Jarvis, 1976, 1982; Jackson, 1976a); however, most channel-bar and channel-fill sequences fine upward. The occurrence of different types of vertical sequence of lithofacies (as shown in Figure 3) does not depend on channel pattern (i.e., degree of braiding or channel sinuosity), but rather on the mode of channel migration, cutting, and filling.

Upper bar deposits [accretionary-bank deposits of Bluck (1971)] and lower bar deposits can be distinguished in modern rivers by their differences in grain size and sedimentary structure, and by the more common presence of buried vegetation in upper bar deposits (Figures 2, 3). Upper bar deposits commonly increase in thickness in the down-bar direction, whereas lower bar deposits decrease in thickness. Such distinctions between upper bar and lower bar deposits have been recorded in ancient examples (e.g., Puidefábregas and Van Vliet, 1978; Van der Meulen, 1982; Diaz-Molina, 1993; Gibling and Rust, 1993). To complicate interpretation, upper bar deposits observed in cores and wireline-logs are easily confused with near-channel overbank deposits. Similarly, in tide-influenced channels, the upper bar deposits tend to contain relatively large amounts of mud and can be confused with the deposits of muddy intertidal flats or coastal bays (Allen, 1991; Dalrymple et al., 1992). This has serious implications for estimates of paleochannel depths and sinuosities. If upper bar deposits are assigned to overbank environments, paleochannel depth and sinuosity are underestimated.

Large-scale inclined strata as shown in Figure 2 rarely have such systematic inclinations, and both discontinuities and discordances are common. Discontinuities in inclination may be associated with the occurrence of unit bars (as discussed in a following section) or with transitions from lower bar deposits to upper bar deposits (Figure 3). Large-scale, lower bar strata may have relatively low inclinations if associated with lower bar platforms (e.g., Bluck, 1971; Campbell and Hendry, 1987; Ikeda, 1989; Bridge et al., 1995). Discordances in large-scale inclined strata form in modern river-bar deposits through discharge fluctuations and shifts in channel position, and (as discussed in following paragraphs) are related to the formation of cross-bar channels (Bridge and Jarvis, 1976, 1982; Campbell and Hendry, 1987; Bridge et al., 1995). Such discordances are recorded in many ancient deposits seen in outcrop (Beutner et al., 1967; Allen and Friend,

Figure 2—Simplified plan forms, cross profiles in various positions and orientations, and large-scale stratal geometries of channel belts for braided and unbraided sinuous rivers. Braid bars and point bars migrated downstream and expanded laterally. Large-scale inclined strata are shown schematically on cross profiles, and upper bar and lower bar deposits are distinguished. The total thickness of bar deposits varies laterally and is generally less than the maximum bankfull depth.



1968; Elliott, 1976; Puidefábregas and Van Vliet, 1978; Galloway, 1981; Van der Meulen, 1982; Thomas et al., 1987; Hirst, 1989; Diaz-Molina, 1993; Gibling and Rust, 1993; Willis, 1993a, b).

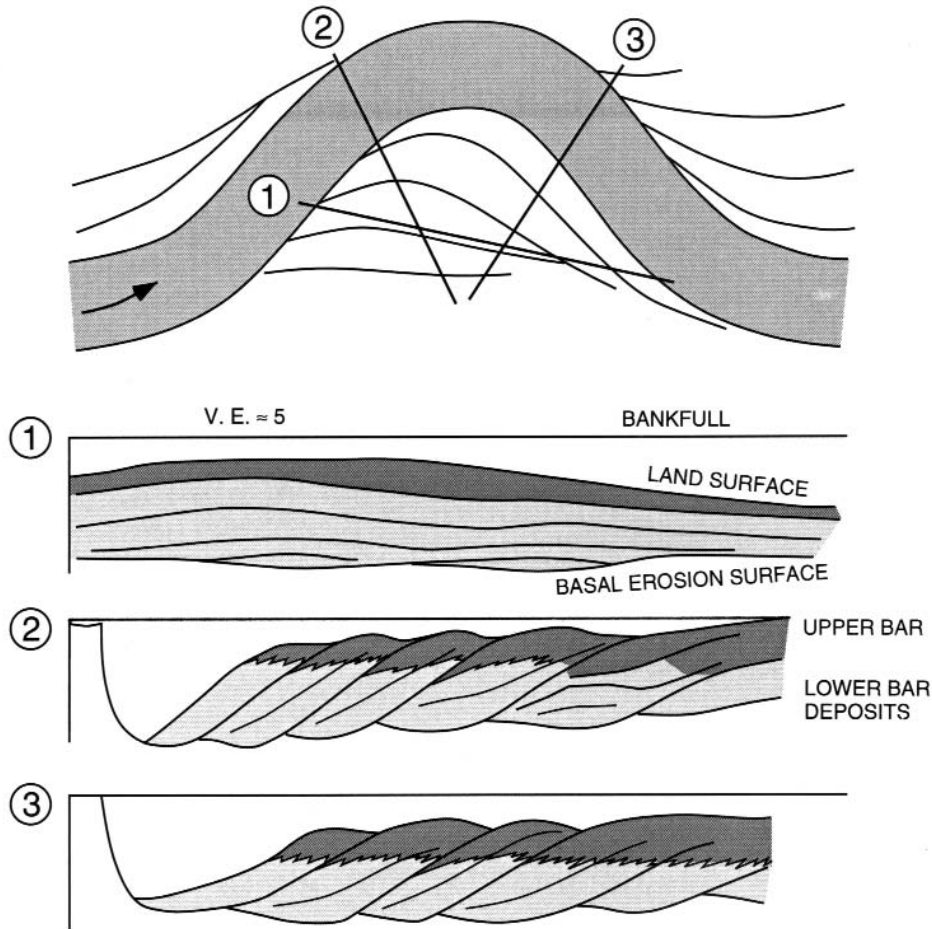
Unit Bars and Cross-Bar Channels

If seasonal deposition is relatively slow or continuous, the large-scale inclined strata are more-or-less sheetlike; however, if channel-bank erosion is sufficiently rapid and intermittent, seasonal deposition may occur on accretionary banks as distinct unit bars (Smith, 1974, 1978; Ashmore, 1982, 1991), that is, as bar-head lobes (Bluck, 1971, 1976; Lewin, 1976) and bar-tail scrolls (Sundborg, 1956; Nilsson and Martvall,

1972; Jackson, 1976b; Bridge and Jarvis, 1982; Nanson, 1980; Bridge et al., 1995) (Figure 4).

Bank erosion and unit-bar deposition change bed- and water-surface gradients. This, in turn, may lead to the formation of new channels cut across braid bars and point bars (Figure 4). Some of these channels develop as the flow takes advantage of the low areas between adjacent bar-head lobes or the “slough” areas between adjacent bar-tail scrolls. Cross-bar channels commonly develop their own bars, the geometry of which is controlled by the flow and sediment transport conditions in these channels. Where a cross-bar channel joins another channel, solitary deltalike deposits with avalanche faces commonly form (e.g., chute bars, tributary mouth bars) (Collinson, 1970;

Figure 2—Continued.



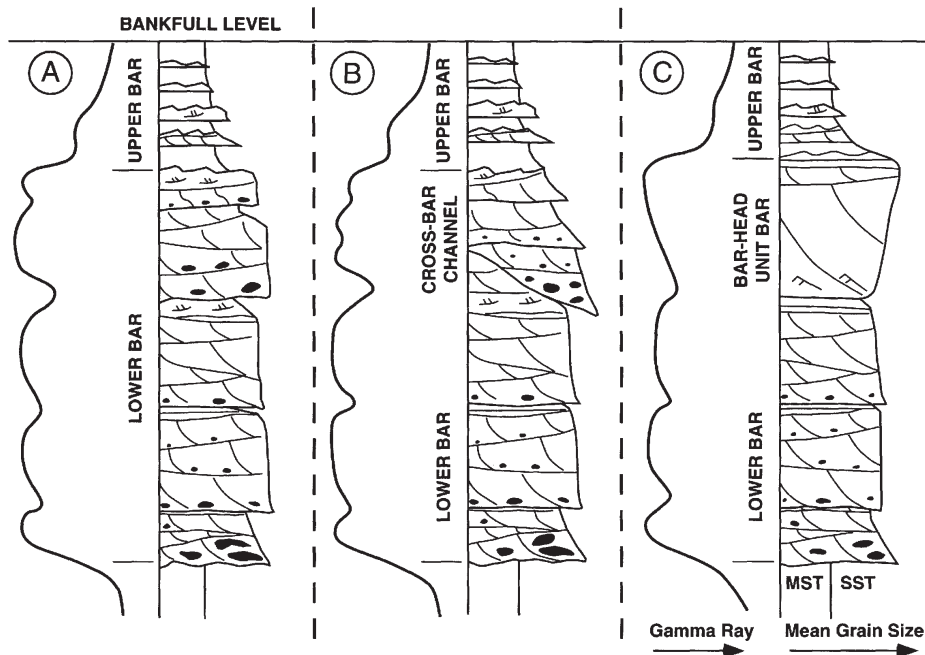
McGowen and Garner, 1970; Smith, 1974; Bluck, 1976; Lewin, 1976, 1978; Levey, 1978; Gustavson, 1978; Cant, 1978; Ashmore, 1982; Ferguson and Werritty, 1983). A cross-bar channel may be enlarged progressively at the expense of an adjacent channel that ultimately is abandoned, thus bringing about changes in the location of main-channel segments. Chute cutoff is an example of such behavior. In other cases, one channel bounding a braid bar may become enlarged at the expense of the adjacent channel.

It is important to try to recognize the deposits of cross-bar channels and their associated unit bars. Failure to do so leads to erroneous estimates of paleo-channel depths and, indeed, of the nature of the fluvial system. Although every gradation in size between a cross-bar channel and a main channel exists, cross-bar channels typically attain a maximum bankfull depth of perhaps one-third or one-quarter of the maximum bankfull depth of the main channel. Thus, if a relatively thick channel-bar deposit has its top truncated by one that is markedly thinner, the uppermost bar deposit may well be associated with a cross-bar

channel (e.g., Figure 3, profile B). Unit bars having avalanche faces will give rise to solitary sets of cross-strata easily confused with those formed by dunes. An exceptionally thick, isolated set of cross-strata, constituting most of the thickness of a large-scale stratum, may represent a unit bar (i.e., bar-head lobe or scroll bar) (Figure 3, profiles C, E), a chute bar (Figure 3, profile E), or a tributary-mouth bar (Figure 3, profile F). Such isolated cross-strata sets have been observed on the upper parts of bars by Collinson (1970), McGowen and Garner (1970), Jackson (1976b), Bluck (1971, 1976, 1979), Cant and Walker (1978), Levey (1978), Blodgett and Stanley (1980), and Crowley (1983). Bar-head and scroll-bar cross-sets tend to occur near the top of the overall bar sequence, whereas tributary-mouth bar (riffle) sets occur nearer the base. If deposition is associated with downstream migration of alternate bars with avalanche faces in straight channels, most of the channel deposits will be composed of a single set of cross-strata (e.g., Smith, 1970, 1971, 1972; Blodgett and Stanley, 1980; Crowley, 1983).

Figure 3—Typical vertical sequences of lithofacies from different parts of sandy channel bars and channel fills. Idealized gamma-ray logs also given.

UPSTREAM PARTS OF CHANNEL BARS



DOWNSTREAM PARTS OF CHANNEL BARS

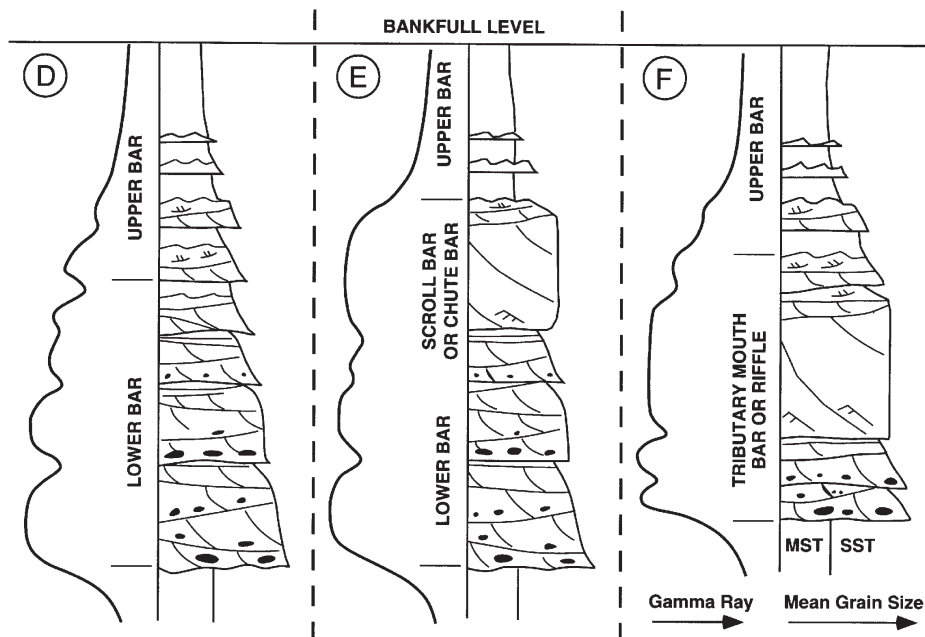
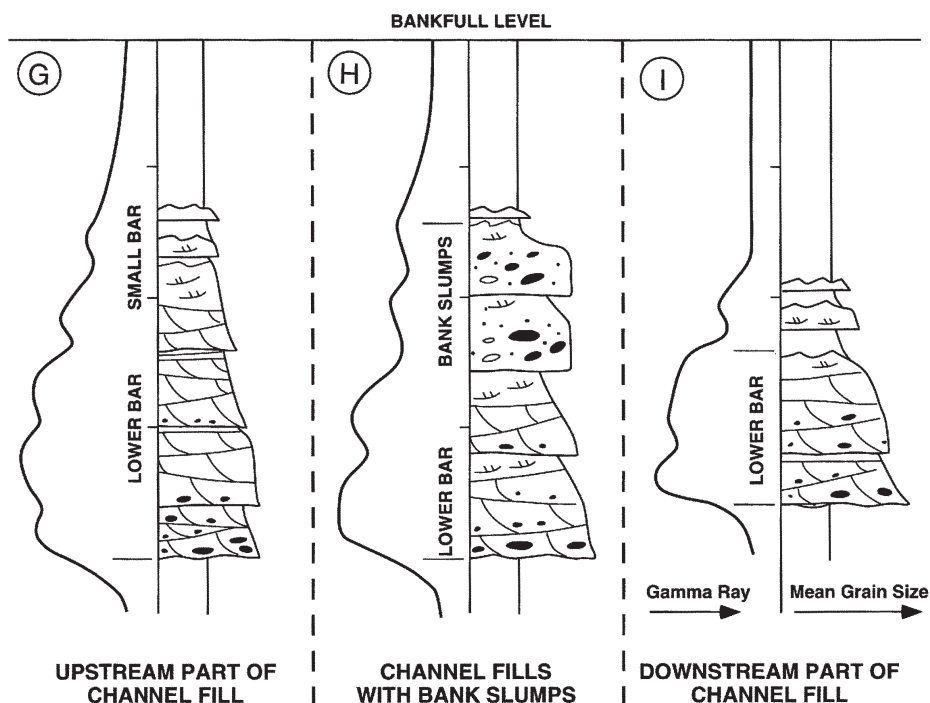
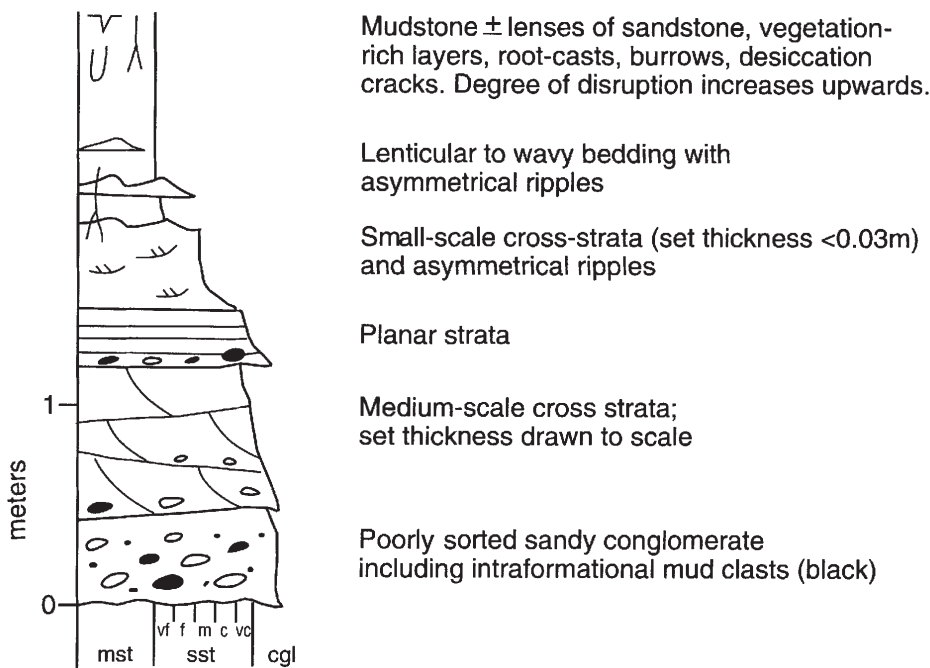


Figure 3—Continued.



LEGEND FOR SEDIMENTOLOGICAL LOGS



Channel-Fill Deposits

The lithofacies preserved within channel fills are dependent on the history of flow through the channels following abandonment. If the angle between

the enlarging- and filling-channel segment is relatively small, as in low-sinuosity rivers, flow is only gradually reduced in the filling channel so that bed load can be deposited, particularly at the channel entrance.

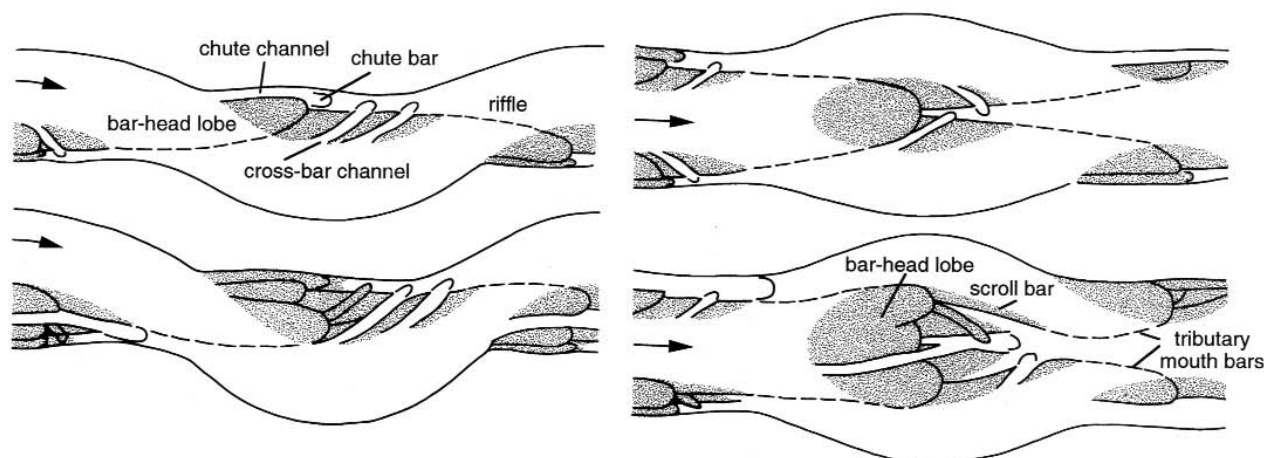


Figure 4—Definition of various types of unit bars and cross-bar channels in idealized braided and unbraided rivers (redrawn from Bridge, 1993, used by permission of The Geological Society).

Although bed load can be transported a considerable distance into abandoned channels, their downstream ends will receive mainly fine-grained, suspended sediment and organic matter from slowly flowing water (e.g., Fisk, 1947; Teisseyre, 1977; Bridge et al., 1986). As angles of divergence increase, both ends of the abandoned channel are quickly blocked so that most of the channel fill is fine grained and organic rich due to suspension deposition in ponded water (Fisk, 1947).

Channel fills generally fine upward, reflecting progressively weaker flows during filling (e.g., Williams and Rust, 1969; Bridge et al., 1986) (Figure 3, profiles G–I). Channel fills also generally fine downchannel as the upstream channel end is choked by bed-load deposition. The relatively coarse-grained bed-load deposits at the upstream end of the channel fill tend to fine upward because they represent progradation of bar-tail deposits into the channel entrance (e.g., Bridge et al., 1986) (Figure 3, profile G). Bed-load deposits in channel fills may also show evidence of accretion on progressively smaller bars as discharge is reduced (Figure 2) (Fisk, 1952; Schumm, 1977; Bridge et al., 1986, 1998). The deposits of these relatively small bars will have large-scale inclined strata, the dimensions of which will decrease with bar size as the channel is filled (Figure 3, profile G). Small deltas may prograde into entrances of abandoned channels containing ponded water (e.g., Gagliano and Howard, 1984), thereby producing coarsening-upward profiles. Sediment-gravity flows from cut banks may accumulate in thalwegs as poorly sorted, structureless deposits (Bridge et al., 1986) (Figure 3, profile H). The suspended-load deposits drape over existing bed topography. Horizontal suspended-load

deposits commonly onlap inclined channel margins. In humid climates, peat may accumulate in the ponded water of channel fills where suspended sediment loads are low (Kosters and Bailey, 1983). In arid climates, evaporitic tufas may form.

Channel-fill deposits grade laterally into channel-bar deposits, and it may be very difficult to distinguish these facies associations, particularly in the subsurface. Channel-fill sequences can look very similar to channel-bar tail deposits. The deposits of the relatively small bars within channel fills may look similar to the deposits within cross-bar channels. The fine-grained parts of channel fills may look very similar to overbank deposits, including lacustrine deposits; nevertheless, accurate estimation of the thickness of channel-fill sequences has implications for calculating paleochannel depth and sinuosity of curved channel segments.

Internal Structure of Seasonal Flood Deposits (Large-Scale Strata)

As the large-scale inclined strata within channel-bar and channel-fill sequences represent episodic, seasonal deposition during floods, they can be recognized by vertical changes in grain size and sedimentary structure (Figure 3). Commonly, large-scale inclined strata fine upward, at least near their tops. The internal structure of large-scale inclined strata in most sandy and gravelly channel-bar deposits is medium-scale cross-strata (sets thicker than about 3 cm, generally an order of magnitude thinner than channel-bar thickness) and planar strata, produced by the migration of dunes and low-relief bed waves. Small-scale cross-strata (set thickness less than 3 cm), arising from migration of

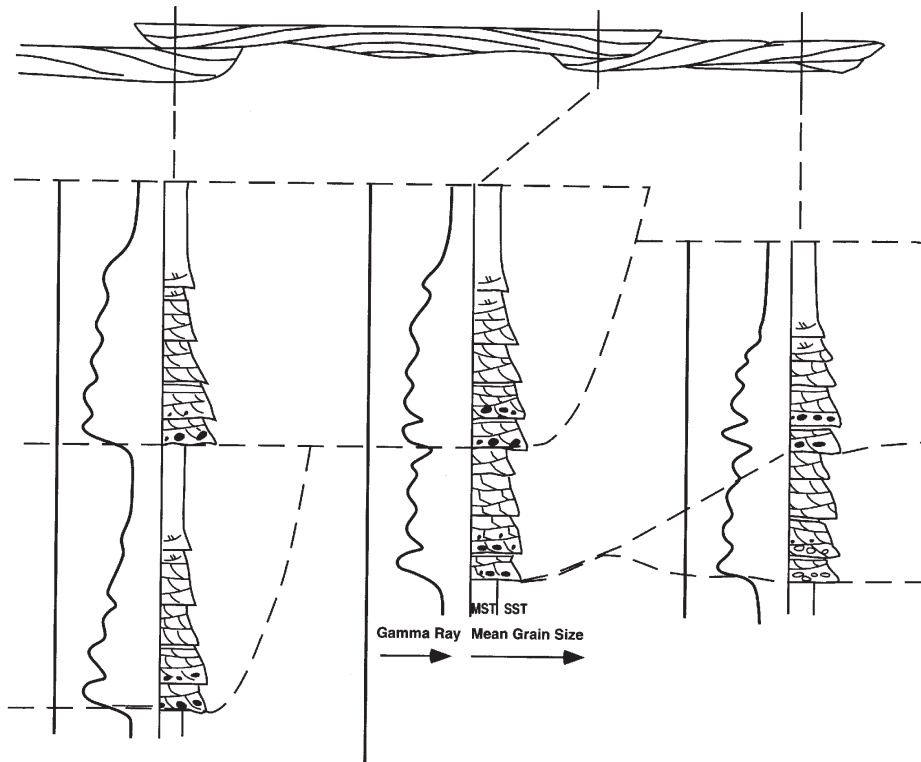


Figure 5—Vertical sequences of lithofacies and gamma-ray logs for superimposed channel bars and channel belts.

current ripples, are normally limited to upper bar deposits and channel fills. In cores, it may be difficult to distinguish medium-scale cross-sets formed by dunes from those formed by some types of solitary unit bars.

Superimposed Channels and Channel Belts

Distinguishing superimposed channel bars and fills within a single channel belt from superimposed channel belts can be very difficult using core and wireline-log data (Figure 5). The ability to make this distinction hinges on the ability to recognize the different superimposed scales of strata set previously discussed. No interpretation problem occurs if easily recognizable overbank deposits (e.g., paleosols) separate the channel-belt deposits. Superimposed channel-bar or channel-fill sequences (e.g., fining-upward large-scale strata sets) having similar thickness (i.e., with little erosional truncation at the top) were probably deposited in separate channel belts; however, the possibility of an unusually deep cross-bar channel cut into a main-channel bar cannot be ruled out. If a complete channel-bar deposit overlies a much thinner one, it is very difficult to discern whether the lower, truncated channel bar or fill deposit was formed in the same or a different channel

belt from the thicker one (Figure 5). Truncation of a relatively thick channel-bar deposit by a deposit that is markedly thinner may indicate that the uppermost bar deposit was formed within a cross-bar channel.

In view of the uncertainties in distinguishing the deposits of single and superimposed channel belts, other kinds of evidence must be considered. Theoretical models of alluvial architecture predict that channel-belt superposition is unlikely if the proportion of channel deposits (net-to-gross) is less than about 0.4. Superimposed channel belts are a certainty if channel-deposit proportion exceeds about 0.75 (Bridge and Mackey, 1993b); furthermore, examination of thickness ratios for the different scales of strata set in channel-belt deposits may shed light on the problem, as discussed in the following section.

METHODS FOR INTERPRETING PALEOCHANNEL DEPTH FROM CORES AND WIRELINE-LOGS

Estimating the width of subsurface channels and channel belts requires confident estimation of maximum bankfull channel depth (see Figures 2, 3). Maximum bankfull channel depth is normally

estimated from the thickness (decompacted) of complete, untruncated channel-bar or channel-fill sequences interpreted from wireline-log or core data. Correct estimation of maximum bankfull channel depth is not always straightforward because complete channel-bar or channel-fill sequences may be difficult to identify, and the thickness of the sandstone-conglomerate parts of these coarse members is not always as great as the bankfull channel depth (Figures 2, 3). The presence of sandy-muddy upper bar deposits, and the uncertainty in distinguishing these deposits from near-channel overbank deposits, makes it difficult to identify paleobankfull level. In addition, within a single channel belt, maximum channel depth and bar thickness can vary spatially by at least a factor of two (Figures 2, 5); therefore, data limited to a single well may not be representative.

An independent means of estimating bankfull flow depth is clearly required. This is possible using a relationship between the distributions of dune height and set thickness of medium-scale cross-strata (Leclair et al., 1997; Bridge, 1997; S. F. Leclair, 2000, personal communication) and the known relationship between dune height and water depth. If using this method, it must be accepted that the distribution of cross-set thickness is determined primarily by variability in dune heights, and that variation in aggradation rate plays a minor role (justification in Leclair et al., 1997; Bridge, 1997); furthermore, application of this method is limited to homogeneous cosets of cross-strata (i.e., no obvious spatial changes in cross-strata type or mean grain size), implying that the cross-strata were formed by migration of dunes with mean geometry that did not vary appreciably in time and space.

To use this method, the thickness, s , of as many cross sets as possible should be measured, such that the mean, s_m , and standard deviation, s_{sd} , of set thickness can be calculated. An initial test of the applicability of the method is that s_{sd}/s_m should approximately equal 0.88 (± 0.3). If so, mean dune height, h_m , can be estimated (S. F. Leclair, 2000, personal communication) as

$$h_m = \alpha\beta \quad (1a)$$

$$\beta \cong s_m/1.8 \quad (1b)$$

$$\alpha \cong 4-8 \quad (1c)$$

or

$$h_m = 2.22\beta^{1.32} \quad (2)$$

Use of equations 2 and 1b is preferable because it does not require estimation of parameter α . To avoid confusing medium-scale cross-strata of dune origin with solitary sets formed by unit bars, abnormally thick, isolated cross-sets should be avoided. Because dune height is expected to vary with position on channel bars, cross-set thickness measured in different positions in the vertical profile should be grouped into subsets.

Mean dune height generally increases with formative flow depth, although the scatter is large (e.g., Allen, 1984). Most of the data fall within the range of $3 < d/h_m < 20$ (d = flow depth). The reason for the large scatter is due primarily to two reasons. First, the height of dunes relative to their length and flow depth increases from near zero at the lower boundary of their hydraulic stability field (transition from ripples or lower stage plane beds) to a maximum in the middle of the field, and then to near zero at the transition to upper stage plane beds. Second, when measurements of dune height and flow depth are made in natural rivers, it is not certain that the dunes were in equilibrium with the prevailing flow conditions. For equilibrium dunes, the maximum steepness, h_m/l , is approximately 0.055 (l = dune wavelength), l/d is approximately 6, and the minimum d/h_m is approximately 3. There are a number of empirical relationships between mean dune height and depth, for example, from Yalin (1964) (equation 3a) and Allen (1970) (equation 3b):

$$d/h_m = 6 \quad (3a)$$

$$d = 11.6 h_m^{0.84} \quad (3b)$$

$$0.1 \text{ m} < d < 100 \text{ m}$$

It appears that, for all types of river dunes (including those not in equilibrium with the flow), d/h_m averages between 6 and 10. Although estimation of flow depth from dune height is imprecise, such an estimate is still a useful complement to flow depth calculated from channel-bar thickness.

To illustrate the use of this method, core data from Mississippi River point-bar deposits can be used (Jordan and Pryor, 1992). Cores 1 and 2 (Figure 6) were obtained from just upstream and just downstream of a channel-bend apex, respectively. Core 1 shows little vertical variation in grain size, but actually fines upward, then coarsens upward slightly. Core 2 fines upward slightly. These vertical trends in grain size are expected based on the core locations (e.g., Bridge and Jarvis, 1976, 1982; Bridge et al., 1995; Jackson, 1976a). The thickness of the point-bar sequences, hence the maximum bankfull channel depth, is 22–24 m. Mean bankfull

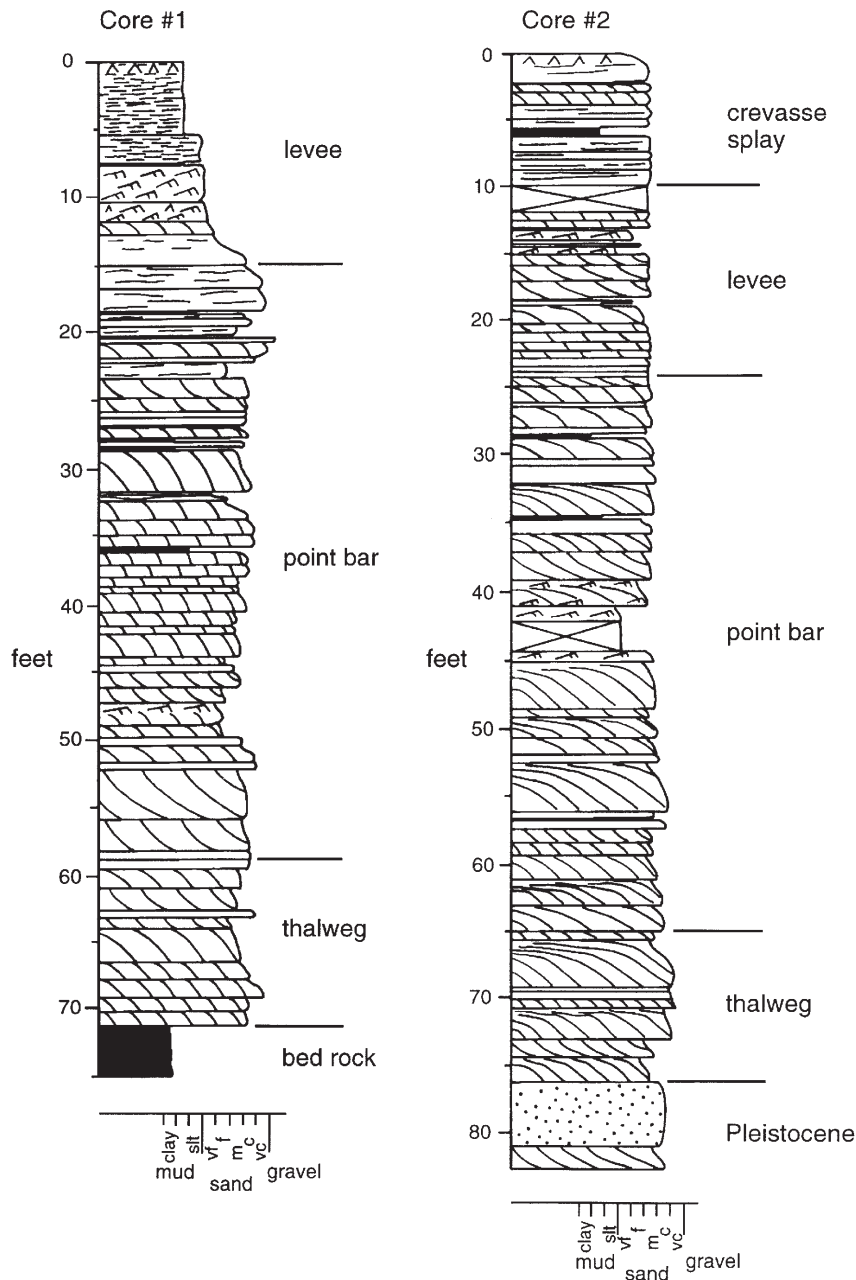


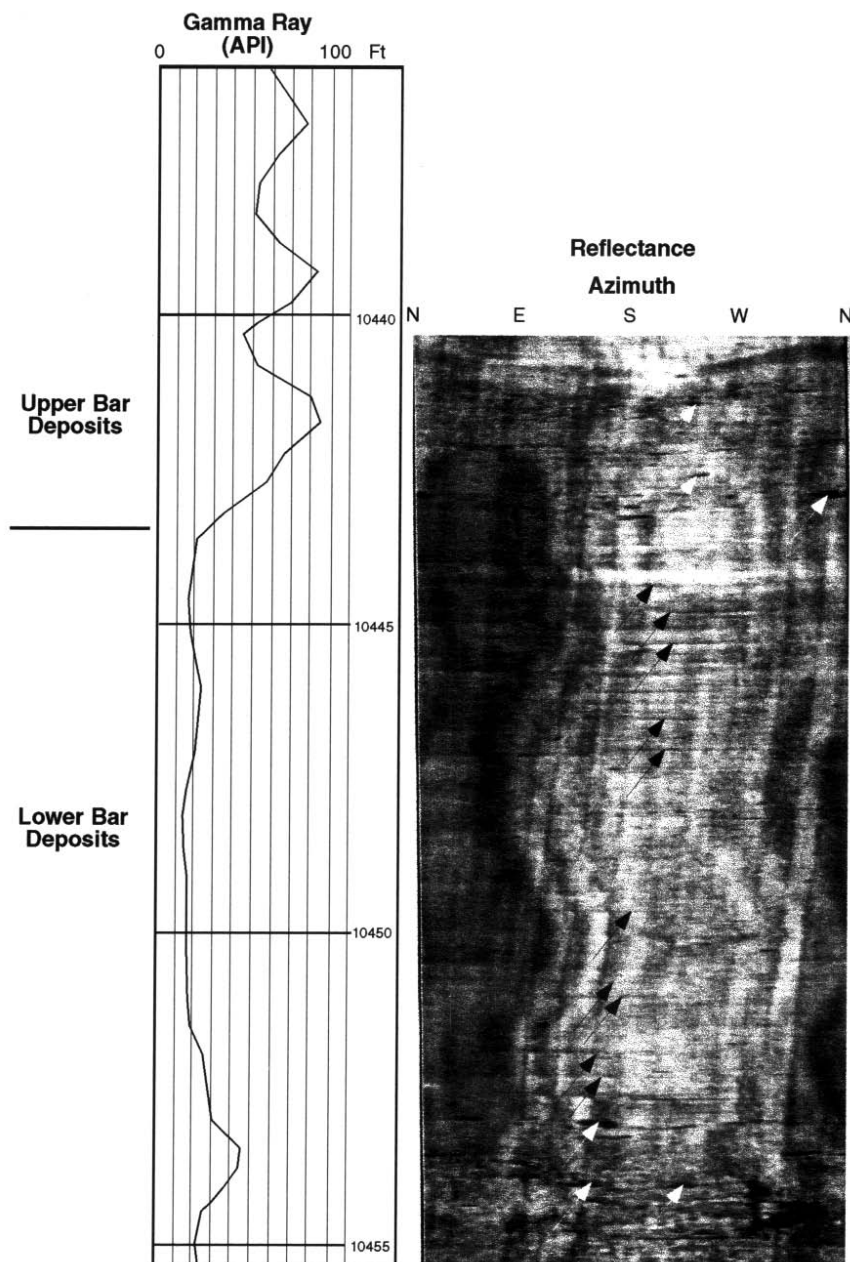
Figure 6—Sedimentological logs of cores from a Mississippi River point bar (redrawn from Jordan and Pryor, 1992).

channel depth is approximately one-half of the maximum bankfull depth, i.e., 11 m.

Cross-set thickness in the cores is quite variable, and there is a slight decrease upward in mean set thickness. In the lower one-half of the point-bar sequences, mean set thickness is 0.42 m and standard deviation is at least 0.27 m (underestimated because the thinnest sets could not be measured accurately). In the upper one-half of the point-bar sequences, mean set thickness is 0.34 m and standard deviation

is at least 0.22 m. Using equations 1b and 2, mean height of dunes responsible for these cross-strata is 1.07–1.42 m. These reconstructed mean dune heights agree closely with those reported by Jordan and Pryor (1992) of 0.9–1.52 m. The mean cross-set thickness and the predicted mean dune height in the central section of the cores are 0.38 and 1.24 m, respectively. If it is assumed that the dunes and cross-strata formed during bankfull flow conditions, the mean bankfull flow depth/mean dune

Figure 7—Gamma-ray and acoustic-imaging logs through interpreted river channel-bar deposits (sandstone and pebbly sandstone) from the North Slope of Alaska. Some cross-set boundaries are indicated by black arrows. Cross-set thickness measured from the acoustic log was used to estimate dune height and flow depth (see text for discussion). White arrows denote claystone clasts. The acoustic log was calibrated with core data.



height is approximately 8.9, falling within the common range of 6–10.

Another illustration of this method involves use of acoustic-image logs and gamma-ray logs (Figure 7). In this example, cross-set thickness in a channel-bar deposit was interpreted from acoustic-image logs calibrated to cores. Cross-set thickness averaged over the channel-bar thickness was 0.16 m, yielding a mean dune height of 0.55 m using

equation 1 with $\alpha = 6$. Flow depth associated with this mean dune height is likely to range from 3.3 to 5.5 m. Maximum bankfull flow depth estimated from the gamma-ray log is about 5.3 m., close to the maximum depth estimated from cross-set thickness. Thus, image logs can be used to estimate cross-set thickness, dune height, and flow depth, although cores will give more accurate results.

METHODS FOR ESTIMATING WIDTH OF SINGLE AND MULTIPLE CHANNEL BELTS

Single Channel Belts

Four commonly used methods for estimating the geometry of isolated channel belts are (1) measurement of outcrop analogs, (2) well-to-well correlation, (3) using empirical equations relating maximum channel depth, channel width, and channel-belt width, and (4) amplitude analysis of 3-D seismic horizon slices.

Outcrop Analogs

The geometry and lithofacies of channel and channel-belt deposits determined in outcrops are commonly used as analogs for subsurface strata (Collinson, 1978; Walderhaug and Mjos, 1991; Lowry and Raheim, 1991; Cuevas Gozalo and Martinus, 1993; Dreyer, 1993; Dreyer et al., 1993; Robinson and McCabe, 1997). Despite the popularity of this approach, it has many pitfalls. First, it must be established that the depositional setting interpreted for the subsurface strata is indeed analogous to that interpreted for the outcrops. This requires two interpretation steps, the reliability of which depends on the quality of the outcrops, of the subsurface data, and of the depositional models used during interpretation. It is difficult to make detailed interpretations of depositional environments using typical subsurface data. Because our understanding of modern depositional environments is incomplete, most depositional models are qualitative, lacking in detail, and not fully 3-D (Bridge, 1985, 1993). The deficiency of most depositional models severely limits their use in detailed interpretation of ancient deposits. Rarely are outcrops extensive enough to allow unambiguous determination of the 3-D geometry and orientation of channels and channel belts (exceptions include Willis, 1993a, b; Zaleha, 1997; Khan et al., 1997; Bridge et al., 2000). Channel-belt width is particularly difficult to determine (Geehan and Underwood, 1993); moreover, channel-belt width and thickness are known to vary spatially within one channel belt and between different channel belts. In general, limited data from a few large exposures are unlikely to be generally representative of fluvial-deltaic channel belts. This is why it is desirable to use analog data from Holocene depositional environments, where channel-belt dimensions can be determined easily, and the relationship between the nature of the deposits and the geometry, flow, and sedimentary processes of the environment can be established unambiguously.

Well-to-Well Correlation

Correlation of specific channel-belt sandstone bodies using wireline-logs is the most common

method for estimating channel-belt widths and orientations. The spatial resolution of this technique can be no better than the average well spacing; therefore, if well spacing is 300 m, sandstone bodies less than 300 m wide cannot be resolved. The validity of this technique is very much dependent on the correlation rules used. Once a suitable horizontal datum has been chosen for the wells to be correlated, it is necessary to establish whether sandstone bodies at similar stratigraphic levels in different wells can be correlated. Well-to-well correlation is commonly compromised by simplistic or erroneous assumptions, such as (1) basal erosion surfaces and tops of channel-belt sandstone bodies are flat, (2) sandstone bodies positioned at the same stratigraphic level must be connected between adjacent wells, (3) sandstone-body width/thickness ratios are closely related to paleochannel pattern, and (4) vertical sequences through channel deposits indicate the paleochannel pattern and hence the geometry of channel-belt sandstone bodies.

Concerning assumption 1, the depositional models discussed clearly show that the basal erosion surfaces and tops of channel belts are not generally flat. Salter (1993) in particular emphasizes the degree of relief of basal erosion surfaces. Concerning assumption 2, sandstone bodies at the same stratigraphic level in adjacent wells are not necessarily connected, and some assessment of the probability of connection is required, perhaps using empirical data on channel-belt width/maximum channel depth; however, if two sandstone bodies are indeed continuous between wells, they are not necessarily from a single channel belt. This is of particular concern if sandstone-body proportion exceeds 0.4 (Bridge and Mackey, 1993b). Concerning assumptions 3 and 4, observations of many modern and ancient channel belts give estimates of channel-belt width/maximum channel depth of between about 700 and 20 (Bridge and Mackey, 1993b). It is commonly stated that this ratio will be larger for braided rivers than for meandering rivers. This is a moot point when using only core and wireline-log data because such a distinction between paleochannel patterns cannot be made reliably; furthermore, this supposition is not generally correct. For example, the channel-belt width/maximum bankfull depth for the meandering lower Mississippi River approximates that of the braided Brahmaputra River, ranging from about 250 to 700 (Coleman, 1969; Bristow, 1987; Fisk, 1947; Jordan and Pryor, 1992; Bridge, 1999). Although channel-belt width is undoubtedly controlled by the channel sinuosity and by the degree of channel splitting, it also depends on bank resistance and the life-span (avulsion periodicity) of the channel belt. These influences are not well understood.

Empirical Equations

Several attempts have been made to predict channel-belt width in the subsurface using empirical equations (derived from modern rivers) that relate maximum channel depth, channel width, and channel-belt width (Collinson, 1978; Lorenz et al., 1985, 1991; Fielding and Crane, 1987; discussed by Bridge and Mackey, 1993b). This approach requires reliable estimates of maximum bankfull channel depth from one-dimensional subsurface data. As discussed, correct estimation of maximum bankfull channel depth is not straightforward because complete channel-bar or channel-fill sequences may be difficult to identify, and the thickness of the sandstone-conglomerate parts of these coarse members is not always as great as the bankfull channel depth; furthermore, within a single channel belt, maximum channel depth and bar thickness can vary spatially by a factor of at least 2 (Bridge, 1993; Salter, 1993). Limited data from a single well thus may not be representative.

Empirical equations relating maximum channel depth, channel width, and channel-belt width have a lot of scatter and are dependent on channel-pattern parameters, such as channel-bend wavelength and sinuosity (Bridge and Mackey, 1993b). Most of the empirical equations used to date (Collinson, 1978; Lorenz et al., 1985; Fielding and Crane, 1987) do not include such dependencies. Channel-bend wavelength and sinuosity are actually very difficult to reconstruct from outcrops and are impossible to determine from well data. Lorenz et al. (1985) were able to apply this technique using empirical equations derived from rivers with sinuosity greater than 1.7 and assuming that sandstone bodies in both outcrops and subsurface strata were deposited in highly sinuous rivers. Equations presented in Bridge and Mackey (1993b) are more generally valid because they are based on broader data sets than previous equations or on theoretical principles. Some of these equations were tested successfully in an outcrop study by Bridge et al. (2000).

Amplitude Analysis of 3-D Seismic Horizon Slices

Amplitude analysis of 3-D seismic horizon slices (Weber, 1993; Hardage et al., 1994, 1996; Burnett, 1996) is the only method capable of directly yielding the width of channel belts and imaging the channel pattern (sinuosity, channel splitting) of subsurface sandstone bodies. This is also the only method that can be used to predict the spatial distribution of channel-belt thickness and lithofacies. These are major advances; however, the integrity of 3-D analyses depends on the resolution of the seismic data relative to the thickness of the sandstone bodies imaged, and requires calibration by wireline-logs and

cores. In general, sandstone-body thickness must be greater than approximately 10 m.

Superimposed Channel Belts

Bridge and Mackey (1993b) used their revised 2-D (two-dimensional) model of alluvial architecture to study the width and thickness of sandstone bodies representing single or connected channel belts. The dimensions of such sandstone bodies depend on the proportion and degree of connectedness of channel-belt deposits in the cross section. For low values of channel-deposit proportion (less than about 0.4), channel belts are unconnected, sandstone-body width equals channel-belt width, and sandstone-body thickness equals aggraded channel-belt thickness. As channel-deposit proportion increases, some channel belts become connected, the mean and standard deviation of width and thickness increase, and their frequency distributions become polymodal with the largest modes at the lowest values of width and thickness. As channel-deposit proportion continues to increase, the distributions are still polymodal, but the larger modes are at higher values of width and thickness. If channel-deposit proportion exceeds about 0.75, all channel belts are connected, and the single sandstone body has a width equal to flood-plain width and a thickness equivalent to the whole section. Channel-deposit proportion, sandstone-body width, and sandstone-body thickness increase as bankfull channel depth and channel-belt width increase and as flood-plain width decreases. Channel-deposit proportion and sandstone-body dimensions also increase as aggradation rate and avulsion period decrease. Alluvial architecture is also influenced by tectonic tilting of the flood plain. For example, channel-deposit proportion and connectedness of channel belts increase on the downtilted side of flood plains, but are reduced on the uptilted side.

The 3-D model of alluvial architecture of Mackey and Bridge (1995) gives results that are generally similar to those described; however, the 3-D model predicts that channel-deposit proportion and connectedness and the dimensions of sandstone bodies vary with distance from points of channel-belt splitting (due to avulsion). Upvalley of avulsion points, sandstone bodies have lower than average width/thickness because of aggradation in a fixed channel belt. Immediately downvalley from avulsion points, channel belts are connected, resulting in sandstone bodies with higher than average width/thickness. Relationships derived from 2-D models between channel-deposit proportion and connectedness and sandstone-body dimensions are strictly applicable only to

parts of the flood plain located some distance downvalley from avulsion points.

A REEVALUATION OF PREVIOUS SUBSURFACE INTERPRETATIONS

There are many examples in the literature of interpretation of river-channel deposits using data from cores and wireline-logs (e.g., various papers in the volumes edited by Barwis et al., 1990; Ethridge and Flores, 1981; Ethridge et al., 1987; Galloway and Hobday, 1996; Lomando and Harris, 1988; Miall and Tyler, 1991; Selley, 1996). In most examples, a range of possible interpretations of wireline-logs is not considered, and the cores are not described in sufficient detail to allow use of the methods proposed here. In particular, the thickness of medium-scale cross-sets is rarely presented; however, data presented by Lorenz et al. (1991) and Tye (1991), plus some of our own unpublished data, can be used to demonstrate how application of these new techniques has an impact upon interpretations of paleochannel depths, channel-belt widths, and sandstone-body dimensions and connectedness.

Piceance Creek Basin, Colorado

Lorenz et al. (1985, 1991) used empirical equations relating maximum channel depth, channel width, and meander-belt width to predict the width of channel-belt sandstone reservoirs in the Mesaverde Group, Piceance Creek basin, Colorado. In applying this method, maximum paleochannel depth was taken as the thickness of untruncated point-bar deposits (sandstone bodies), decompacted by an arbitrary 10%. Meander-belt widths were calculated to be 350–520 m. These predictions were tested using well-to-well correlations, pressure and tracer tests, and vertical seismic profiling. Figure 8 shows an example of one of the sandstone bodies studied by Lorenz et al. (1991). This sandstone body was interpreted by Lorenz et al. (1991) as composed of three stacked point-bar sequences (see Figure 8). We have reinterpreted this sandstone body (using the concepts discussed) as a single channel bar-fining-upward sequence, resulting in an increase in the interpreted maximum bankfull depth (i.e., to 7.5 m). To assess this reinterpretation, cross-set thickness was used to estimate dune height and flow depth. The mean cross-set thickness in the lower part of the sandstone body is 0.29 m, giving a mean dune height of about 0.87 m. If the flow depth/mean dune height is between 6 and 10, the flow depth corresponding to this part of the sandstone body must range between 5.2 and 8.7 m,

in agreement with the flow depth reconstructed from point-bar thickness. If a maximum paleochannel depth of 7.5 m is used in the empirical equations used by Lorenz et al. (1985, 1991), meander-belt width is estimated to be 1184 m, much wider than originally predicted.

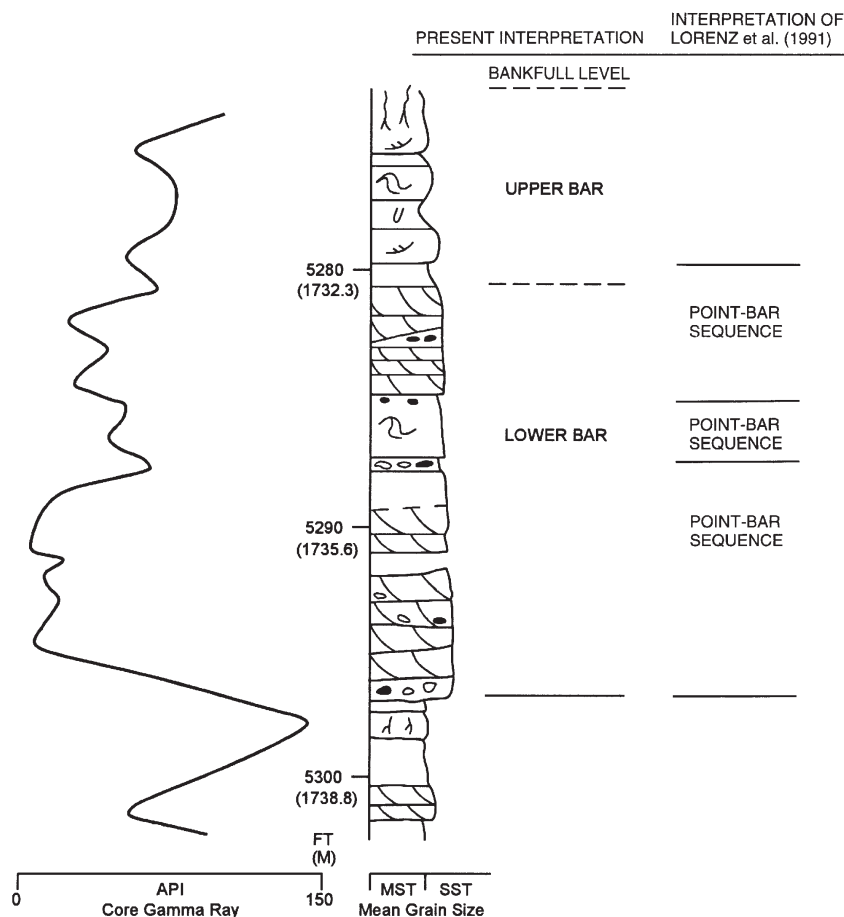
North Appleby Field, Texas

In his study of North Appleby field, Texas, Tye (1991) described the heterogeneity of low-permeability fluvial reservoirs using 36 wells (drilled on 640 ac or 259 ha spacing) and 168 m of core from two wells. Although the lithofacies, descriptions and basic interpretation of depositional environments are sound, the stratigraphic interpretation of the reservoir can be considerably improved and quantified through utilization of methods introduced here.

Wireline-logs and well-to-well correlations of a stratigraphic interval (zone 1) of the Travis Peak Formation discussed by Tye (1991; see also Davies et al., 1993) are shown in Figure 9. Sandstone bodies interpreted as deposits of braided-to-meandering channel belts and flood-plain sandstone bodies were correlated using some of the unjustifiable assumptions discussed previously. The cross sections depict channel-belt sandstone bodies in zone 1 ranging from 2.4 to 8.7 m thick (mean = 6 m). Using thickness/width ratios from supposed modern analogs in combination with the well-to-well correlations, it was determined that the widths of channel belts (cut normal to paleovalley direction) ranged from 4.8 to 9.6 km. Tye (1991) thus concluded that a channel-belt sandstone could be penetrated by as many as 20 wells drilled with 640-ac (259 ha) spacing, implying that nearly two-thirds of the wells in the North Appleby field could be in horizontal communication. Davies et al. (1993) concurred with Tye's (1991) stratigraphic interpretation.

Using the methodology discussed previously, the Travis Peak core and wireline-log data were reevaluated to estimate maximum bankfull channel depths from which channel-belt widths were calculated using empirical equations (Bridge and Mackey, 1993b). The first step in this reevaluation was to assess the interpretation of depositional environment. It is now realized that paleochannel patterns, such as braiding and meandering, cannot be determined reliably from core and wireline-log data. Fining-upward and blocky-sandstone bodies were interpreted by Tye (1991) as channel deposits, whereas interbedded sandstone and mudstone were interpreted as flood-plain and abandoned-channel deposits. Some of these relatively fine-grained deposits may be upper bar deposits (Figure 10). In the section of core shown in Figure 10,

Figure 8—Example of sedimentological and wireline-logs from the Mesaverde Group, Piceance Creek basin (redrawn and modified from Lorenz et al., 1991), with revised interpretation.



maximum bankfull channel depth is reinterpreted as approximately 7 m. Mean cross-set thickness in the lowest 3 m of this sandstone body is 0.24 m, yielding a mean dune height of approximately 0.68 m. If the flow depth/mean dune height is 6–10, the local bankfull flow depth associated with these cross-sets and dunes is 4.1–6.8 m. This estimate agrees with the maximum bankfull flow depth of 7 m reconstructed from the thickness of the channel-bar sequence (Figure 10).

If maximum bankfull flow depth in the Travis Peak Formation ranges from 6 to 10 m, mean bankfull flow depth is 3–5 m, and the range of channel-belt width (*cbw*) is predicted to be 436–1741 m using empirical equations from Bridge and Mackey (1993b):

$$cbw = 59.9 d_m^{1.8} \quad (4a)$$

$$cbw = 192 d_m^{1.37} \quad (4b)$$

Although such empirical equations typically have large standard errors, the predicted mean

channel-belt widths are still considerably less than those originally suggested by Tye (1991). The overestimation of channel-belt sandstone-body width was due mainly to the fact that the correlated sandstone bodies probably constituted a series of connected channel belts.

The revised channel-belt dimensions were used in the 2-D alluvial stratigraphy model of Bridge and Mackey (1993a) to assess the likelihood of connection between individual channel-belt sandstone bodies. It is realized that such a model will not represent the 3-D distribution of sandstone bodies. A model simulation for zone 1 in North Appleby field is shown in Figure 9A. Input parameters are as follows: mean bankfull channel depth is 4.0 m; mean width and standard deviation of channel belts are 730 and 300 m, respectively; mean avulsion period is 300 yr; and mean channel-belt aggradation rate equals 0.01 m/yr. The avulsion period and aggradation rate were selected within realistic limits to simulate the observed channel-deposit proportion (net-to-gross) of 0.49 and channel-belt connectedness ratio of 0.35.

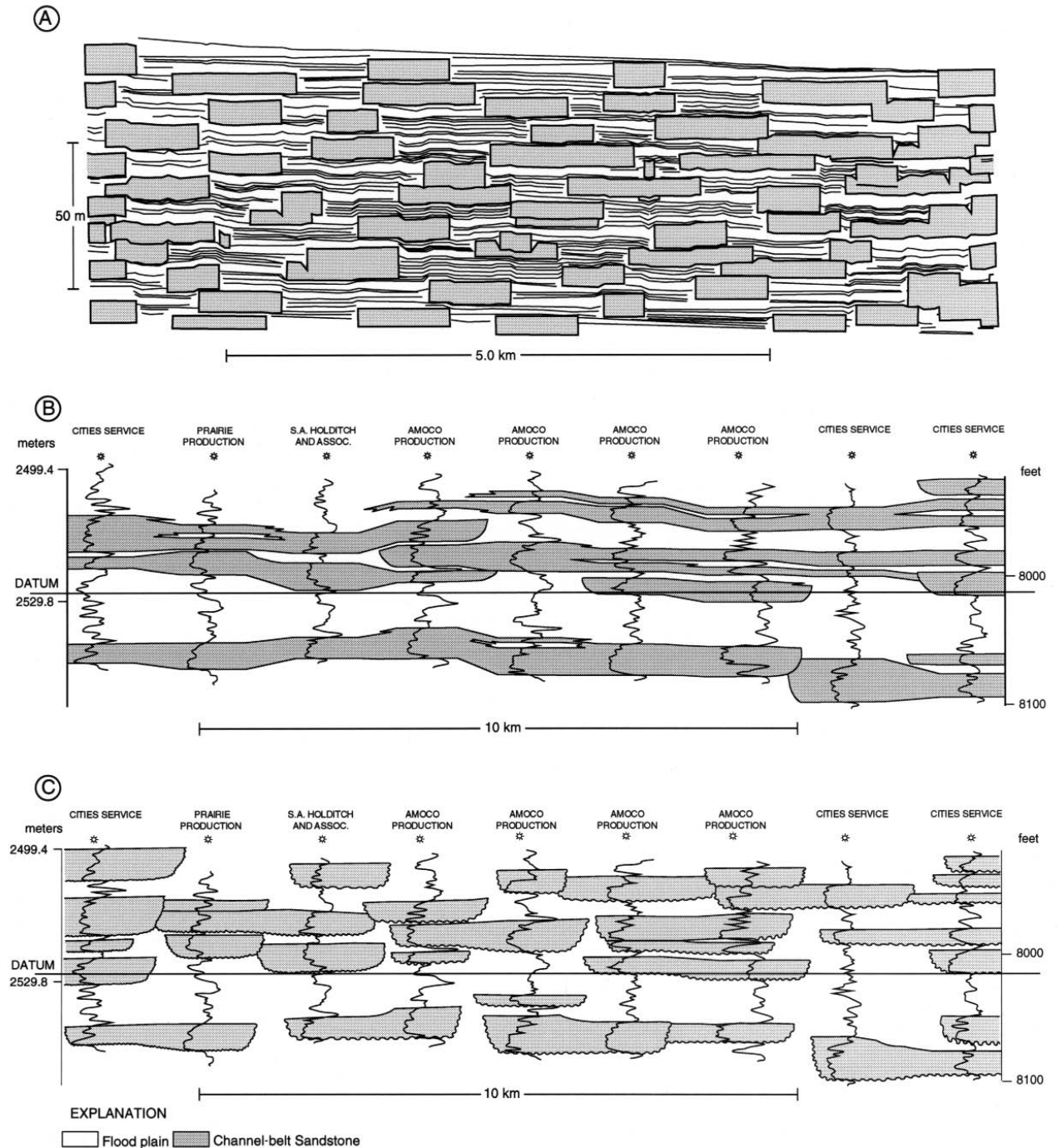
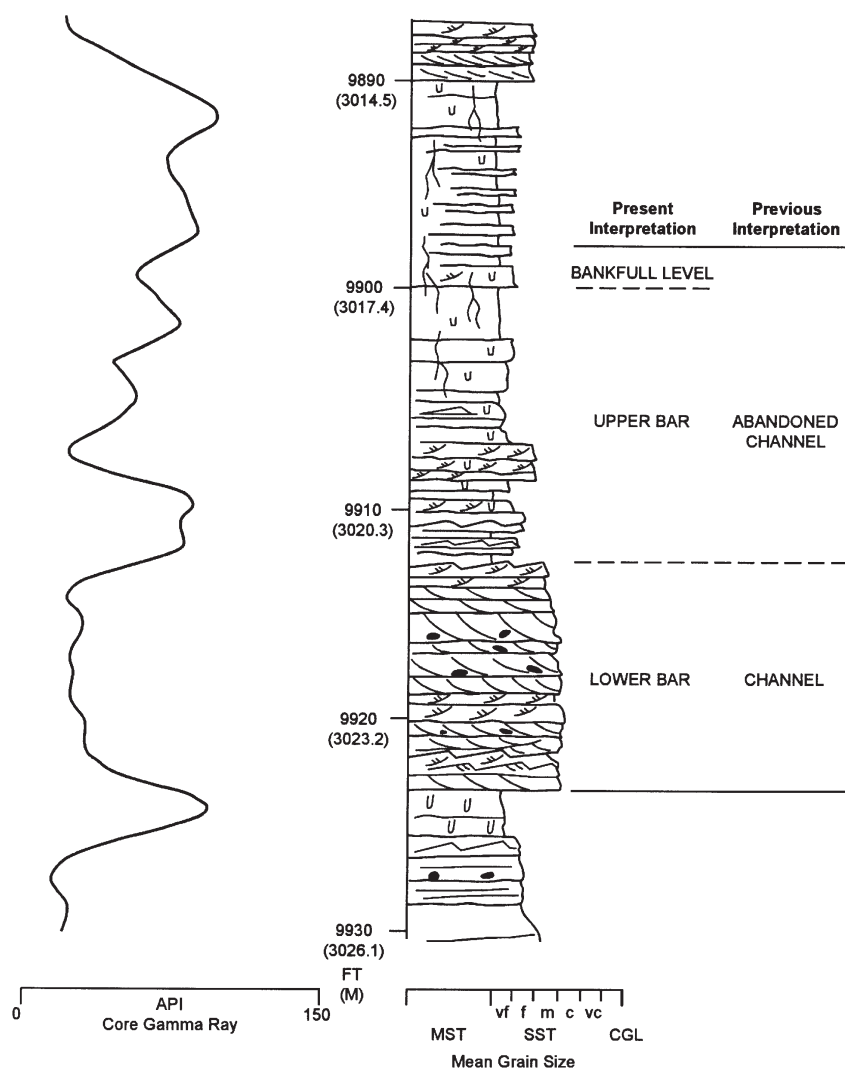


Figure 9—(A) Two-dimensional stratigraphic simulation of fluvial channel belts (stippled) in the Travis Peak Formation, North Appleby field, East Texas basin (zone 1 of Tye, 1991; Davies et al., 1993). (B) Wireline logs and channel-belt sandstone bodies of zone 1 as originally correlated by Tye (1991). (C) Revised well-to-well correlations using same data as in (B), but constrained by recalculated channel-belt widths. Note that limiting channel-belt widths decreased sandstone body continuity in two dimensions.

Figure 10—Sedimentological and wireline-logs from the Travis Peak Formation, North Appleby field, with revised interpretation (redrawn and modified from Tye, 1991). Recognition and proper interpretation of bar-top deposits are critical to accurate estimation of bankfull channel depth and channel-belt width. Bankfull channel depth can also be estimated from the mean and standard deviation of cross-set thickness.



The simulated cross section depicts channel belts that are much narrower and more uniformly distributed than those interpreted from the original wireline-log correlations of Tye (1991) (Figure 9B). Revised correlation (Figure 9C) is based on revised channel-belt thickness and width, and on the spatial distribution and connectedness of the channel belts suggested by the 2-D stratigraphic simulations. The number of wells required to deplete the reservoir in Figure 9B may be significantly less than the number needed to maximize rate and total recovery from the reservoir shown in Figure 9C. Admittedly, channel belts appearing disconnected in cross section may be connected out of section, and an estimate of channel-belt connectivity in three dimensions is required (Mackey and Bridge, 1995); however, prediction of the 3-D form and spatial distribution of channel belts is not the point of this paper.

DISCUSSION

Numerous publications are concerned with the application of fluvial depositional models to reservoir characterization (e.g., Campbell, 1976; Cant, 1982; Collinson, 1978; Collinson and Lewin, 1983; Ethridge and Flores, 1981; Flint and Bryant, 1993; Flores et al., 1985; Galloway and Hobday, 1996; Marzo and Pudefábregas, 1993; Miall, 1978; North, 1996; Selley, 1978, 1996; Tillman and Weber, 1987; Walker and James, 1992; Weimer et al., 1985). Use of qualitative, simplistic depositional models by many geologists has led to reservoir descriptions that are framed in terms such as "ribbon" or "blanket" (sheet) sandstone bodies that are related to paleochannel patterns, such as meandering or braided. When pressed by reservoir engineers for information on the dimensions of permeable and impermeable rock bodies,

on the degree of communication between adjacent wells, and on the nature of permeability barriers, geologists commonly cite width/thickness ratios of depositional units associated with different inferred paleochannel patterns, derived from studies of presumed outcrop analogs (Atkinson et al., 1988; Ayers, 1986; Daams et al., 1996; Doyle and Sweet, 1995; Dreyer et al., 1993; Flach and Mossop, 1985; Flores and Hanley, 1984; Miall, 1994; Olsen et al., 1995; Putnam and Oliver, 1980; Richardson et al., 1987a, b; Robinson and McCabe, 1997; and many others). We suggest here that such approaches to reservoir description are intrinsically untenable, and more rigorous, quantitative approaches are required under high-cost/low-profit market conditions.

One outcome of ambiguous stratigraphic interpretation of fluvial strata is that geologists and engineers are more frequently embracing geostatistics to model channels and channel belts (Hatløy, 1993; Hirst et al., 1993; Høimyr et al., 1993; Seifert and O'Meara, 1993; Williams et al., 1993; Tyler et al., 1994; Ballin et al., 1997; Eschard et al., 1998; Hirsche et al., 1998; MacDonald et al., 1998); however, the outcrop data collected to build geostatistical models are commonly suspect, as discussed. Statistical manipulation of measurements from an inappropriate or misinterpreted outcrop analog produces a quantitative, but flawed, model.

The original and reinterpreted cross sections of North Appleby field contrast two approaches to subsurface stratigraphic interpretation. One approach is limited by qualitative, simplistic facies models, whereas the other approach combines a more sophisticated understanding of fluvial facies with quantitative relationships between the dimensions of dunes, bars, channels, channel belts, connected channel belts, and their associated strata sets. Whereas no subsurface interpretation can ever be assumed to be correct, the latter approach to stratigraphic interpretation is much more likely to approach reality.

Using these improved methods of quantitative estimation of channel-belt dimensions and connectedness with a discovery well and a core through the target zone in the delineation well, many reservoir parameters can be estimated earlier and more accurately than heretofore possible. At the field discovery-appraisal stage, our approach to correlating channel-belt reservoirs can be used to evaluate pressure-support mechanisms (aquifer or gas cap), guide the number and placement of development wells, and influence well locations, type, and completions; furthermore, precise production forecasts and knowledge of probable fluid-handling needs protects against overplanning or underplanning of drilling and production facilities. Finally, as production wells are drilled, a dynamic reservoir model can be built prior to implementation of secondary

and tertiary recovery processes. If interpretation of the dimensions and connectedness of channel-belt sandstone bodies in a reservoir is inaccurate, the resultant permeability field is erroneous no matter how well lithofacies and petrophysical data are calibrated. This point is most often driven home by poorer than expected well performances, unanticipated fluid (water or gas) production, and disappointing performance of secondary and tertiary recovery projects. Thus, many mature fields that did not achieve their initial production estimates required costly infill drilling, perforation modifications, and well conversions (Green et al., 1996; Ambrose et al., 1997; Hohn et al., 1997; Montgomery, 1997; Potocki et al., 1997; Hamilton et al., 1998).

The type, dimensions, and spatial distribution of shales in fluvial reservoirs have a major impact on permeability variations. Presently, the best approach to estimating the dimensions of lacustrine or flood-plain shales is to correlate them in detail (aided by biostratigraphy and geochemistry) and develop a plausible regional stratigraphic interpretation (Eschard et al., 1998; MacDonald et al., 1998). Armed with wireline-log and core data and our methods of predicting bankfull channel depth, channel width, and channel-belt width, the distribution and dimensions of various shale types in reservoirs can be modeled more effectively. Dimensions of abandoned-channel clay plugs can be approximated using calculations of paleochannel depth, width, sinuosity, and wavelength (Bridge and Mackey, 1993b; Bridge et al., 2000). Clay-clast conglomerates at the bases of channel-belt sandstone bodies form significant barriers or baffles to flow (Jones and Hartley, 1993; Doyle and Sweet, 1995; North and Taylor, 1996). The lengths of clay-clast conglomerate layers can be estimated by calculating channel-belt widths using the methods discussed here.

CONCLUSIONS

A fresh approach to quantitative determination of the thickness and width of subsurface channel-belt sandstone-conglomerate bodies is made using recent theoretical, experimental, and field studies. This new approach involves (1) new models for the three-dimensional (3-D) variation of lithofacies and petrophysical-log response of river-channel deposits with explicit recognition of the different superimposed scales of strata, (2) distinction between single and superimposed channel bars, channels, and channel belts, (3) interpretation of maximum paleochannel depth from the thickness of channel bars and from the thickness of sets of cross-strata formed by dunes, and (4) evaluation of methods for estimation of widths of sandstone-conglomerate bodies

that represent either single or connected channel belts (outcrop analogs; correlation of sandstone-conglomerate bodies between wells; use of empirical equations relating channel depth, channel width, and channel-belt width; theoretical models; and 3-D seismic data). Examples of the use of this new approach demonstrate the potential for considerable improvement in quantitative stratigraphic interpretation of fluvial reservoirs and aquifers. This, in turn, should have an important impact on discovery appraisal, reservoir characterization, and management.

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